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SHEAR WAVE VELOCITIES INFERRED FROM SURFACE WAVE DISPERSION BENEATH THE PŘÍBRAM ARRAY IN THE CZECH REPUBLIC

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ABSTRACT

Rayleigh waves in the period range 0.2 – 3.0 s from eight quarry blasts are analyzed to obtain S-wave velocity model beneath the Příbram seven-station array in the Czech Republic. Locations and origin times of blasts are estimated using P- and S-wave onsets and then verified at the quarry in the vicinity of the location. This blind test confirms a sufficient precision of the location procedure for identification of quarries. Epicentral distances are in the range from 16 to 52 km. Group velocity dispersion curves of Rayleigh waves are determined by the frequency-time analysis. An average group velocity beneath the array for each period is computed with the help of mean travel-time curve for all blasts and stations. The resultant group velocity dispersion curve is inverted to obtain a 1-D S-wave velocity model using the Isometric method. The results are compared with known geological structure in the area of interest.

KEYWORDS: surface waves, frequency-time analysis, group velocity, dispersion, shear wave velocity, seismic array

1. INTRODUCTION

Příbram region is a historical mining area. Silver and polymetallic ore was excavated here for several centuries. In the 1950's and 1960's an intensive mining of uranium ore took place near Příbram at the contact of the Central Bohemian Pluton and Barrandian metamorphic rocks (mainly siltstones, shales and wackes). The mining openings reached a depth over 2 km. After decline of mining, underground gas storage of cavern type was built at a depth of 950 m in the granitic rocks of the Pluton.

Quarry blasts are intensive sources of short-period surface waves. There is a relatively dense network of quarries in the Central Bohemia around the local Příbram seismic array (Málek and Žanda, 2004). The seismic array was deployed for monitoring of seismicity induced by Příbram-Háje gas storage. Příbram array has been operating since August 1998 and it is situated around the gas storage.

This paper develops a new approach in processing of Rayleigh waves recorded by the local seismic array using blasts from surrounding quarries. This method is based on the averaging of group velocities for each period from all travel-time curves available. From this procedure we obtain an average dispersion curve for gas storage zone (GS average dispersion curve). The velocity model down to a depth of about two kilometers is inverted from the average group velocity dispersion curve of Rayleigh waves. This model of gas storage zone (GS model) is compared with the model of broader area of Central Bohemian Pluton (BP model) inverted from the average dispersion curve from the granitic area of the Pluton (BP average dispersion curve).

2. DATA

The Příbram array consists of seven three component digital seismic stations equipped with Guralp CMG-40T sensors and RUP2004 acquisition system (Štrunc and Brož, 2004). All seismometers are situated on a surface with a flat topography. An average altitude is about 560 m above see level. Sampling frequency of all the stations is 100 Hz.

During the operation of the Příbram array several hundreds of quarry blasts a year are registered. Blasts from eight different quarries with clear surface waves were selected in the period from July 2007 to February 2008. They have epicentral distances up to 52 km and provide sufficient azimuthal coverage, see Fig. 1. The geographical coordinates of quarries and origin times of blasts were first estimated as a blind test using P- and S-wave onsets. We assumed homogeneous velocity model. P-wave arrivals are detected from the vertical component and are mostly well recognizable. S-wave arrivals are detected on the horizontal components. They are often hidden in P-wave coda and hence the S-wave times are less accurate then the P-wave ones. The errors of P- and S-wave arrivals are typically 20 ms (2 samples) and 40 ms (4 samples), respectively.

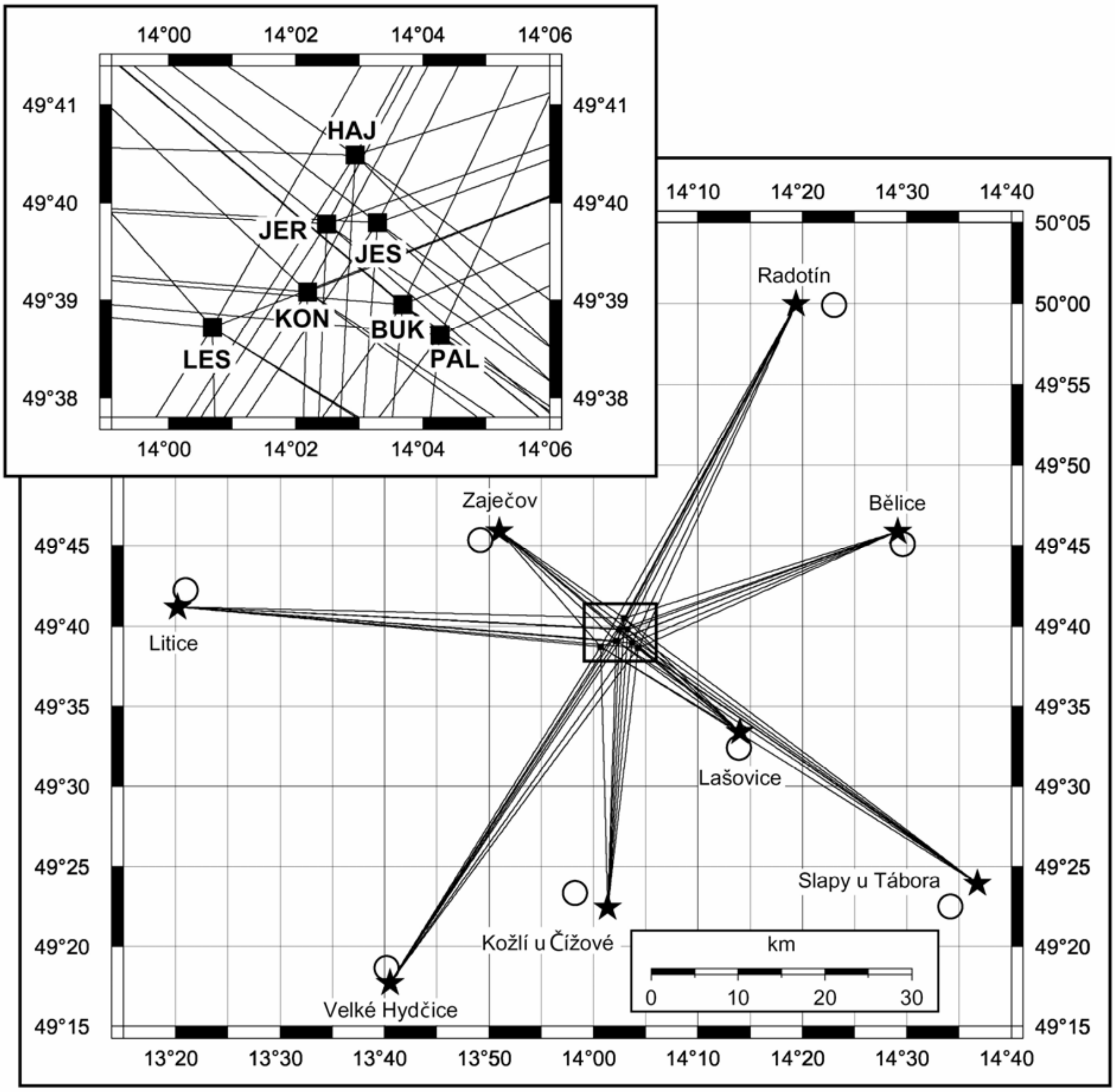

Fig. 1 Map of the seismic stations and used quarry blasts. Stars denote true coordinates of blasts, circles represent computed locations. Seismic stations are represented by squares. Seismic rays are drawn as lines between stations and quarries. Upper insert shows the detail of the Příbram seismic array.

The localization was done in two steps. First, horizontal components of the P-wave slowness vector *p* were calculated. As both quarry blasts and stations are situated on the surface and homogeneous model is used, vertical component is zero. Plane wave approximation is considered. The slowness vector $\boldsymbol{p} = (p_1, p_2)$ can be calculated from the system of linear equations,

$$\sum_{i=1}^{2} \Delta X_{ji} p_i = \Delta t_j \quad , \quad j = 2...7 \,, \qquad (1)$$

where *j* is the station index. The stations are indexed according to their P-wave arrivals. The first station, which serves as the reference station, is the nearest station to the epicenter. $\boldsymbol{\Delta X_j}$ is a position vector between *j*-th and the 1-st station and Δt_j is a P-wave time difference between *j*-th and the 1-st station. Equation (1) represents an overdetermined problem. The optimal solution in L_2 norm is obtained by Gaussian method (Tarantola, 1987).

The direction of vector $\boldsymbol{p}$ determines a backazimuth from the station to the source. Absolute value of the slowness vector determines apparent slowness of P-wave along the surface.

In the second step, epicentral distance is calculated from time differences between P- and S-waves. The epicentral distance R from the particular blast *n* to the central point of the array is estimated according to the formula:

$$R_n = \frac{v_P(t_S^n - t_P^n)}{k-1} \qquad (2)$$

where t_P is the arrival of P-wave, t_S is the arrival of S-wave, v_P is the velocity of P-wave and k is a v_p/v_s ratio.

This method gives only an approximate solution, as the real velocity distribution differs significantly from the homogeneous model. In our case we assumed homogeneous velocity model with v_p of 6.0 km/s and v_s of 3.5 km/s. Subsequently, the nearest quarries were found on the map for the calculated coordinates. The information about the blasts and their exact coordinates were verified directly by the staff of these quarries. Resulting map of the seismic stations and used quarries is shown in Fig. 1. The verified coordinates of the blasts were used for further surface wave processing and interpretation.

The origin time is determined according to a simple formula:

$$t_0 = t_P - \frac{R}{v_p}. \qquad (3)$$

In the case of quarry Lašovice the origin time of the shot have been accurately measured by a technique specially developed for this purpose. Special seismometers BR3 (Brož, 2000) were installed at a distance of tens of meters from the shot to extrapolate the origin time with an accuracy of about 5 ms. For Lašovice shot an accuracy of the above computed origin time was tested. Difference between the computed origin time and the measured one is 0.125 s. The travel time of surface waves was from 5.3 to 8.2 s for periods from 2.4 to 0.3 s. That means that the relative error in the group velocity determination varies from the 1.52 % for the shortest period to the 2.36 % for the longest period. This error however affects only determination of group velocities from the source to the stations. For determination of group velocities between the stations only time differences are important and origin time does not enter computations, see bellow.

Altogether, eight blasts measured at seven stations were used in this study. These 56 seismograms were analyzed using the frequency-time analysis to obtain group velocity dispersion curves.

Usually the first step in the analysis of surface waves is an introduction of instrumental correction. However, for calculation of the GS model only time differences are used, hence the instrumental phase shifts are compensated (all stations are expected to be identical). Also in the case of calculation of the BP model the instrumental corrections can be neglected. Maximum error caused by the instrumental correction is only 0.01 km/s for the period of 2.4 second. For shorter periods this error is even smaller. This value can be neglected with respect to the accuracy of determination of the origin time.

In the present paper we deal with Rayleigh waves and we use only the vertical component of the seismograms.

3. METHODS

3.1. GROUP VELOCITY MEASUREMENT

We look for the group velocity dispersion curve between each source-station pair. To analyze dispersive records, the standard method of Fourier transform-based multiple filtering is applied. The spectrum of record is multiplied by a weighting function centered at many discrete frequencies. Non-constant relative resolution Gaussian filtering is used, for details see Dziewonski et al. (1969). Examples of estimating the optimum coefficient for controlling the width of the filters can be found in Levshin et al. (1972 and 1992). In the present paper, a linear dependence of the width coefficient on a period is used. For details on estimating this dependence see Kolínský (2004).

A result of the multiple filtering is a set of quasimonochromatic signals. Computation of an analytical signal corresponding to each of these signals is also provided. A modulus of the analytical signal represents an envelope of the quasimonochromatic signal. Maxima of the envelopes give the dispersion curve.

An envelope of quasimonochromatic signal has often several local maxima, forming several ridges in spectograms obtained by frequency-time analyses. They belong to different modes of surface waves. The ridge corresponding to fundamental mode is selected in such a manner that the resultant dispersion curve is smooth regardless of the absolute values of the amplitudes involved in this ridge. This approach enables to analyze even the records where body wave amplitudes exceed surface waves. Procedure starts with longer periods where surface waves dominate the record and then it proceeds along the continuous ridge to shorter period range where surface waves may be hidden among other wave-groups. Details on selecting the dispersion ridge are described in Kolínský and Brokešová (2007).

A filtered seismogram is created by summing the truncated quasimonochromatic signals. It contains only the fundamental mode of Rayleigh waves as shown in Figure 2 by dashed lines. These filtered wavegroups are shown to highlight the surface waves in the raw seismogram. For further interpretation only the dispersion curves are used.

3.2. GS AVERAGE DISPERSION CURVE

First step consists of selection of the period range where a majority of group velocity dispersion curves was estimated. Time differences of the arrival times ΔT_j given by the envelope maxima of the corresponding quasimonochromatic signals at all stations j are set for each chosen period. Then the distances ΔR_j between the reference stations with

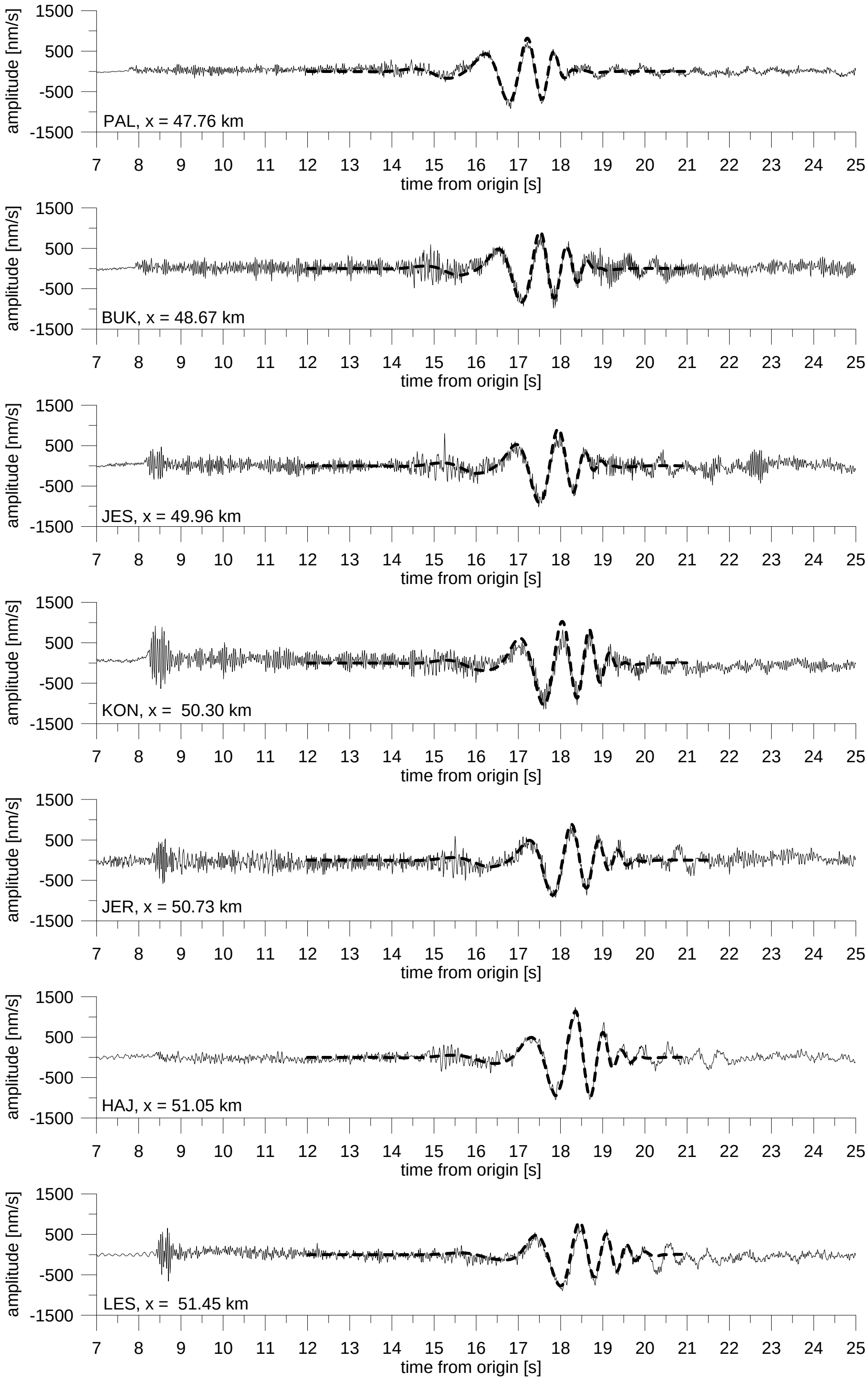

Fig. 2 Example of measured seismograms – vertical components of seismograms from all seismic stations of the Příbram array. The source is situated in the quarry at Slapy u Tábora, date: 31.8.2007, origin time: 11:01:38.39 UTC. Solid lines show measured data, dashed lines represent interpreted surface wavegroup.

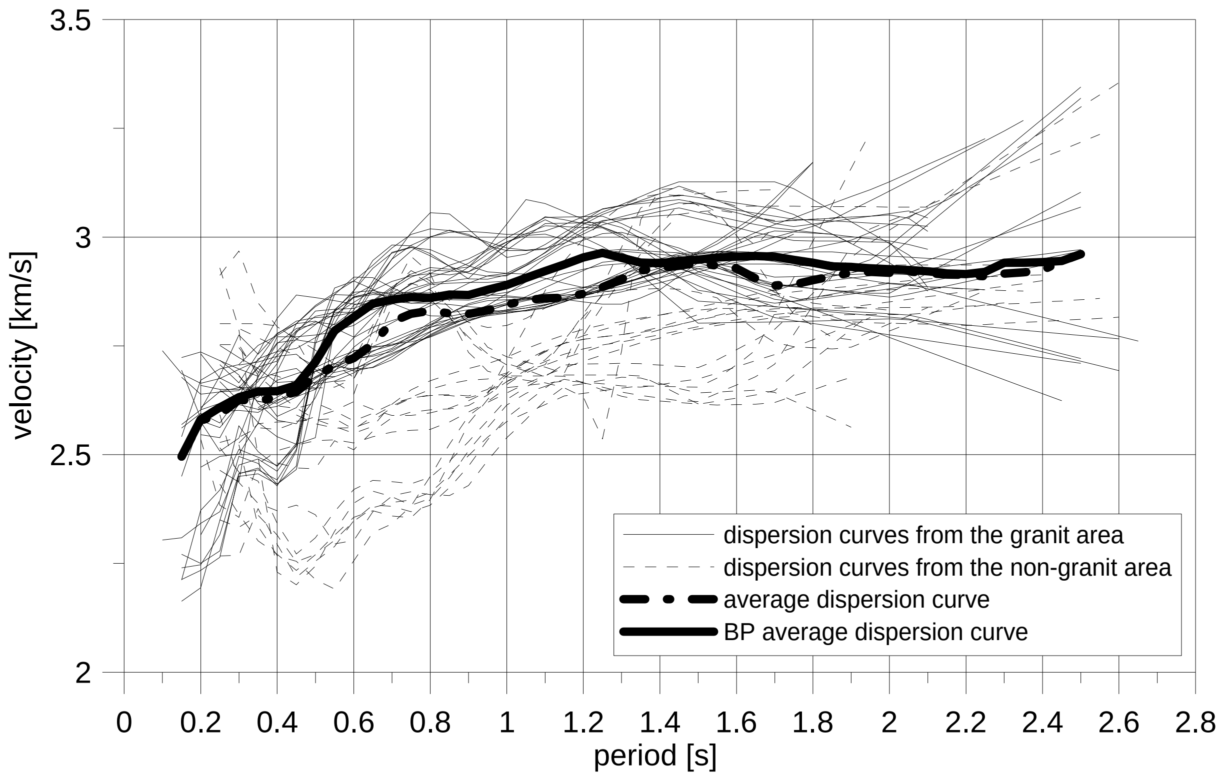

Fig. 3 Measured dispersion curves between the sources and the Příbram array stations.

minimal epicentral distance for a given quarry blast and all other stations are set as a differences of their respective epicentral distances. The average group velocity for each of the chosen period is computed as a slope of a linear regression (type $y = ax$) fitted into the data of ΔR_j and ΔT_j. Obtained values of these group velocities for given periods form the GS average group velocity dispersion curve.

3.3. INVERSION

The applied procedures are described in detail by Kolínský and Brokešová (2007). The Isometric Method (IM) is used, which is a fast inverse algorithm developed by Málek et al. (2005 and 2007). It combines features of several standard methods, particularly the simplex method, Newton's least-squares method and simulated annealing, see Tarantola (1987). Typical problems, which are effectively solved by the IM, are weakly non-linear problems with tens of parameters and complicated forward modeling. Therefore it is quite suitable for the inversion of dispersion curves.

The forward problem – dispersion curve computation – is solved by the modified Thomson-Haskell matrix method; see Proskuryakova et al. (1981). Dispersion curves are computed in a 1-D layered structure above a halfspace with constant values of v_s, v_p and densities in the individual layers and in the halfspace.

During the inversion, the group velocity dispersion curve is computed many times and the deviation between theoretical and measured dispersion points (misfit function) is minimized. The thicknesses of layers are set manually and kept fixed during the inversion. In this study, layers of 0.1 – 0.3 km thicknesses are used, which seems to have a sufficient resolution for finding the properties of the velocity distribution and keeps the number of parameters reasonable.

Since Rayleigh wave group velocity curves are inverted, three parameters in each layer need to be found: v_s, v_p and density. As the dispersion depends predominantly on v_s, the other two parameters are constrained. The density is fixed to increase in constant steps with increasing depth. The value of 2.5 g/cm^3 is considered in the uppermost layer and the density increases by 0.05 g/cm^3 in each next layer.

Seeking for v_p is constrained by the v_p/v_s ratio, which is set to be 1.73 ± 0.1. These values are often considered, see e.g. Novotný and Urban (1988). Given the a priori constrains on v_p and density, we do not present these parameters as a result of the inversion in our study. For details refer to Kolínský and Brokešová (2007), where several tests of the inversion reliability and a discussion on the resolution of the procedure can be found.

4. RESULTS

An example of raw data is shown in Figure 2. There are seismograms of the vertical component from all seven stations of the Příbram array from the quarry blast at Slapy u Tábora. Graphs are arranged according to their epicentral distance. Fundamental mode of Rayleigh waves, which was used for the interpretation, is represented by a dashed line.

Dispersion curves of group velocities were estimated by means of frequency-time analysis from the vertical component of the seismograms. The average dispersion curve (Fig. 3) was calculated as a median value of group velocities from all stations and all sources for each period. There are significant differences among the dispersion curves estimated from blasts in metamorphic rocks (from quarries Litice, Zaječov and Radotín) and the dispersion curves from blasts originated in the granitic rocks of the Central Bohemian Pluton (quarries Bělice, Slapy u Tábora, Lašovice, Kožlí u Čížové and Velké Hydčice). Dispersion curves from granitic area have higher group velocities. This is caused by the differences in shear wave velocities that strongly affect Rayleigh wave group velocities. Shear wave velocities in granitic area were found to be generally higher than those in metamorphic rocks. For better illustration of this phenomenon the average dispersion curve calculated only from the curves from the granitic area (BP average dispersion curve) is shown by bold line in Figure 3.

The GS average dispersion curve was computed from all the measured dispersion curves. Graphs of ΔT versus ΔR together with fitted lines of linear regression are displayed in Figure 4. Period interval was selected from 0.3 s up to 2.4 s with a step of 0.1 s. For each period also standard deviation of the linear regression was calculated. Higher values of standard deviation are in the beginning and in the final part of interpreted period interval. For comparison also the same graphs for P wave and S wave are plotted in Fig. 4. In the case of body waves, arrival times were used instead of times of the envelope amplitude maxima. Inverted velocity for P-wave and S-wave velocities from this method are 6.00 km/s and 3.41 km/s.

We do not obtain the same number of $\Delta T/\Delta R$ pairs for all the periods because some periods are not well recognizable due to the noise and hence various period ranges of dispersion curves were obtained during the frequency-time analysis. GS average dispersion curve was calculated from about 50 values for each period. For longer periods this number decreases to almost a half of the value. This fact is shown in Fig. 4, where significantly lower density of points in the long period graphs is obvious.

Fig. 5 shows the resultant GS average dispersion curve together with the BP average dispersion. Resultant GS average dispersion curve, calculated from the data presented at Figure 4 by the method described in paragraph 3.2, is represented by the points of average group velocity for each interpreted period. For each point a vertical line representing standard deviation of the linear regression is added. Ten inverted best fits are also drawn in Figure 5. An average misfit value between the measured and synthetic dispersion curves normalized by the number of measured dispersion points is 0.05 km/s.

Both GS model and BP model are depicted in Figure 6. In this figure the GS model (velocity model of the gas storage area covered by the array) is compared with the BP model (average velocity model of the Central Bohemia Pluton), which was obtained from group velocities between the quarries in granite and the array. In both models, the uppermost two layers have been estimated using only the shortest surface waves and hence their reliability is lower. The deeper layers of both models differ significantly. The GS model has lower velocities to the depth of 400 m. In the depth interval from 0.9 km to 1.5 km there is a low velocity channel in the GS model.

5. DISCUSSION

The difference between the GS model and the BP model can be partly explained by errors of origin time estimation which is used during BP model computation (see chapter 2). For this reason the GS model is more reliable. It represents S-wave velocity model in the locality of the underground storage, where a lot of mining openings are situated. Old mining tunnels could be responsible for the low velocities down to the depth of 500 m. The low velocity channel at the depth of about 1 km corresponds to the depth, where the cavern of gas storage is situated (950 m). The storage represents 45 km of tunnels, where 620 000 m^3 of gas is stored. Although the wavelengths of surface waves are much longer than the diameter of individual tunnels (diameter about 6 m), the average rock properties can be affected by an existence of the openings.

6. CONCLUSION

In this paper we developed a new method for calculation of the dispersion curves of surface waves at a local array of seismic stations using quarry blasts outside the array. Method is based on the evaluation of the average group velocity for each period from the travel-time curves for all blasts and stations. The S-wave velocity model derived from this average dispersion curve represents the geological structure bellow the seismic array.

The method was applied to the seismic array Příbram-Háje in the locality of the underground gas storage. The obtained S-wave velocity model down to the depth of 1.8 km differs significantly from the average model of the Central Bohemian Pluton. The differences are interpreted as the effect of mining openings and cavern of the gas storage.

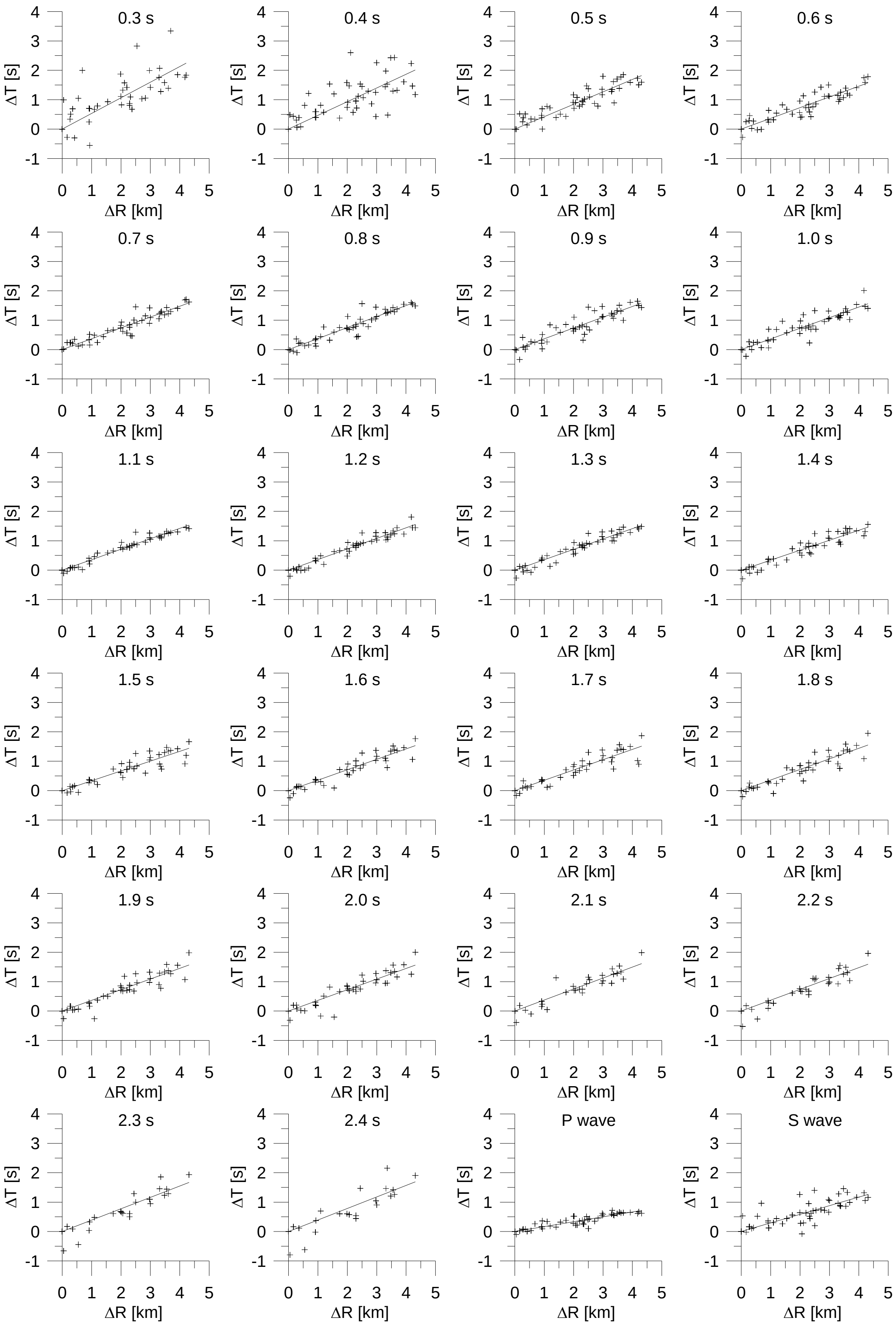

Fig. 4 Graphs of differences between epicentral distances (*ΔR*) and times (*ΔT*) for periods from 0.3 s to 2.4 s. Last two plots show analogous results for P- and S-wave.

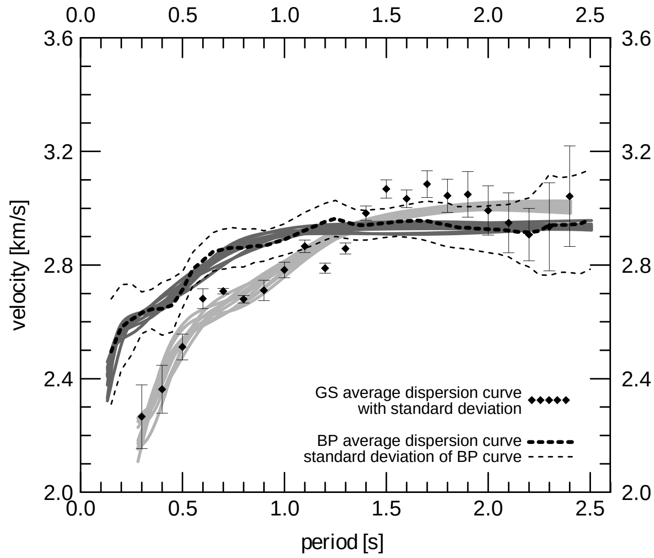

Fig. 5 Fit between GS average dispersion curve (diamonds) and inverted dispersion curves corresponding to found velocity structures (light gray lines) of the GS model. BP average dispersion curve (bold dashed line, the same as bold thick line in Figure 3) and corresponding inverted curves (dark grey lines) for the BP model. Standard deviations are drawn for respective measured curves.

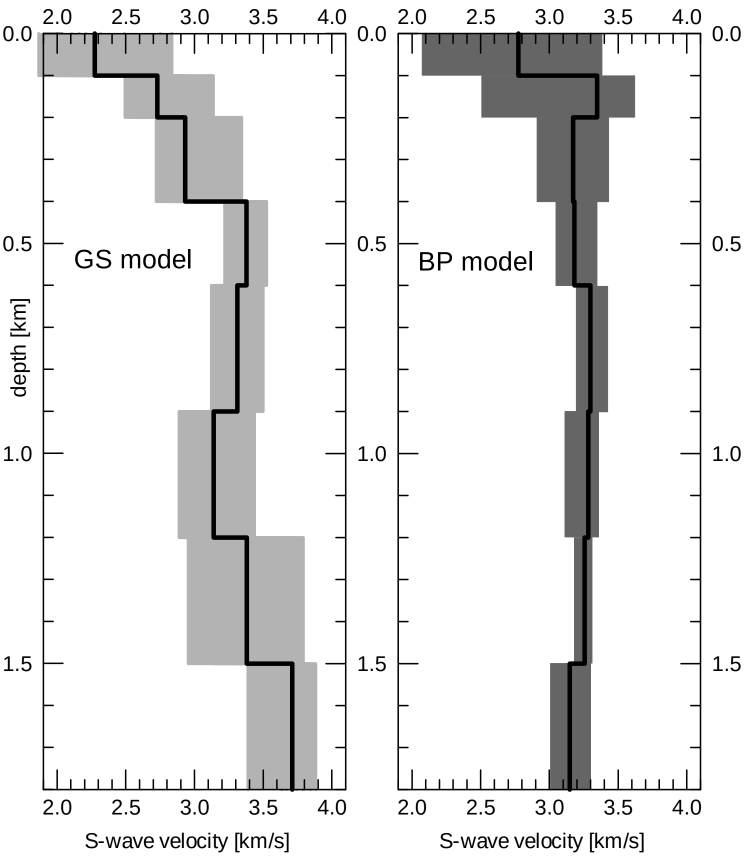

Fig. 6 GS model and BP model. Grey shaded areas represent uncertainty of the inversion and bold solid lines show resultant average shear wave velocity distribution.

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